Understanding of Seismic Activity Using Conductivity Data in the Central Part of Northeastern Japan

Yukio Fujinawa\(^1\), Noriaki Kawakami\(^2\), Jun Inoue\(^2\), Theodore H. Asch\(^2\), Shinji Takasugi\(^3\) and Yoshimori Honkura\(^4\)

\(^1\)National Research Institute for Earth Science and Disaster Prevention, Tennodai 3-1, Tsukuba-shi, Ibaraki 305-0006, Japan  
e-mail: fujinawa@bosai.go.jp

\(^2\)Geothermal Energy Research and Development Co., Ltd., 11-7, Kabuto-cho, Nihonbashi, Chuo-ku, Tokyo 103-0026, Japan

\(^3\)Geothermal Engineering Co., Ltd., 804, Koami-cho, Nihonbashi, Chuo-ku, Tokyo 103-0016, Japan

\(^4\)Tokyo Institute of Technology, Ookayama 2-12-1, Meguro-ku, Tokyo 152-8551, Japan

Abstract. Quasi-3D conductivity distribution in the central part of northeastern Japan Arc is built from results of wide-band (0.002–20,000 Hz) magnetotelluric measurements (MT). The observation has been conducted along three traverses in the central Tohoku district at 86 observation sites in order to better understand seismic activity, tectonics, and geology there. Long MT traverses were supplemented by three short traverses to investigate seismic activity in the Central Mountain Range. We used the MT impedance tensors corrected for the subsurface 3-D galvanic distortion effects including static shift effects.

Modeling results indicate that the crust is vertically homogeneous without enhanced conductive lower crust under the tectonic scenario of the back-arc region. The profiles delineate two near-surface clear conductive anomalies in the fracture zone between the Dewa Hill and the Central Basin, and in the zone between the Kitakami and Abukuma River regions suggesting the role of pore-fluids in weakening of the crust. Conductors in the crust, west of the Central Mountain Range, generally correlate well with mapped faults or pre-Tertiary tectonic lines with high seismicity. The buried fault between the Central Basin and the Central Mountain Range is suggested from the conductivity and earthquake hypocenter data. Several other geophysical parameters such as the Poisson ration are introduced to understand the seismotectonics of a given region.

We illustrated that the conductivity is one of the most important parameters in interpretation of seismotectonics.

1. INTRODUCTION

Knowledge of resistivity distribution in the crust is expected to provide valuable constraints in understanding tectonics, seismicity and metamorphic activity. Combined use of the georesistivity data with other geophysical data, such as gravity, geomagnetism, and elastic parameters enables to reveal the regimes of crust and upper mantle. The Tohoku area is located in a typical subduction zone with active volcanic and seismic activity (Fig. 1). To construct geodynamic model of the island arc, the structure of the crust in the Tohoku has been
Fig. 1. (a) Simplified geological maps of the northeast Japan arc and the three MT profiles. Lines A, B, and C (Fujinawa et al., 1997). Three short transects D, E, F supplement the data to the north of the main survey area. Miocene and Quaternary volcanic fronts are shown with several tectonic lines. The MT transects to the north (Akita-Iwaizumi) and south (Niigata-Abukuma) of the present study area were surveyed by Ogawa (1992). The northernmost parts have been surveyed by Nabetani and his group (e.g., Nabetani et al., 1992). (b) Location of the MT measurement and geologic divisions (After Geological Survey of Japan ed., 1995). Triangles denote Quaternary volcanoes. TTL is the Tanakura Tectonic Line and HTL is Hatagawa Tectonic Line. TEM measurements were also conducted at each site to infer the static-shift.
investigated since 1990, using seismic and electromagnetic approaches (Yokokura et al., 1992). Here, we report the contributions of electromagnetic research, made so far, in this region.

The geological and tectonic setting is briefly described here (see Ogawa (1992) for more detail). The Tohoku district is divided into the Green-Tuff region in the west and the Non-Green Tuff region to the east of the Morioka-Shirakawa Line, running parallel to the roughly north-south Quaternary volcanic front (Figs. 1 and 2). The Green Tuff region has experienced intense volcanism during early Miocene and Quaternary period around the Central Mountain Ranges along the Quaternary volcanic front (VF). Quaternary volcanism gave rise to a chain of volcanoes and an associated geothermal field located at the intersecting region of the present volcanic front and pre-Tertiary tectonic lines. Thick Neogene and Quaternary marine sediments are accumulated in the inter-mountain and back-arc basins.

The Non-Green Tuff region is composed of the Northern Kitakami, the Southern Kitakami and the Abukuma sub-regions, each of which is bounded by the pre-Tertiary tectonic lines of the Hayachine, the Hatagawa and the Tanakura, respectively (Kimura et al., 1991). The basement rocks of the Northern Kitakami Belt are part of the Asian continental convergent margin sequence and are composed of resistive Triassic and Jurassic marine sediments. The Southern Kitakami Belt is accreted to Paleozoic and Mesozoic formations and granites (Saito and Hashimoto, 1982). Late Cretaceous left-lateral strike-slip motion
along the Hatagawa tectonic line juxtaposed the southern Kitakami and Abukuma belts (Taira et al., 1983). The Cretaceous Tanakura tectonic line separates NE-Japan and SW-Japan. In the Tohoku district, the structural trend of the Neogene units is nearly perpendicular to the subduction direction of the Pacific plate. But these units are also affected by the structure of the basement rocks of pre-Tertiary age (Kimura et al., 1991), suggesting 3-D structural configuration.

In northeast Japan, detailed MT surveys have been conducted in the northernmost region (Nabetani et al., 1992; Nabetani and Fukuta, 1993; Nabetani and Maekawa, 1994; Nabetani and Fukuta, 1995a, b; Nabetani and Kimura, 1996), the northern region (Ogawa, 1992), the southern region (Ogawa, 1992), and the central region (Utada, 1987). The present three traverses are situated between the two surveys of Ogawa (1992) and one traverse is the same as that of Utada (1987). However, the present observations are more extensive in frequency and number of observation sites. Another MT transect has been conducted in the north of our survey area (Mishina, 1999). Our data will be combined with data collected through networked-MT array at longer periods (Uyeshima et al., 1994, 1995), providing improved georesistivity imaging in the northeast Japan arc.

We constructed 2-D conductivity models along each transect using the impedance tensors after allowing for static shift and distortion effect caused by sub-surface three dimensional metrogeneities. The static-shift was estimated using TEM (Transient Electromagnetic Measurements) data, which is known to be an efficient method to infer the site gain (Sternberg et al., 1988). The MT tensor distortion was corrected through the GBD procedure (Groom and Bailey, 1989, 1991; Groom et al., 1993), treating them as a galvanic effect.

In northeastern Japan, an extensive seismic network has been established by Tohoku University (e.g., Hasegawa et al., 1991), which has provided detailed information on seismic discontinuities such as the Conrad and Moho (Horiuchi et al., 1982), structure of the subducted Pacific plate (Umino and Hasegawa, 1975), and seismic velocity structure (Hasemi et al., 1984; Obara et al., 1986; Zhao et al., 1992). The structural features inferred by seismic techniques could be more clearly interpreted if a detailed georesistivity structure is obtained, because resistivity is known to be sensitive to pore water and partial melt that to great extent control the seismic processes. Particular attention is directed to conductivity contrast in the upper and lower crust in relation to depth cut-off of seismicity in the upper crust. Seismicity of different zones is discussed in the light of conductivity data as well as other geophysical and geological data.

2. OBSERVATION AND DATA

Here the observational procedure is introduced briefly. Details are given elsewhere (Fujinawa et al., 1997, 1999; Fujinawa, 2000). Broadband magnetotelluric data were acquired at 86 observation sites in the central part of the northeastern Japan arc on three traverses (Lines A, B, C from north to south), each about 150 km long and running approximately east-west from the coastal area of the Pacific Ocean to that of the Japan Sea (Figs. 1 and 2). Five components magnetotelluric fields in the wide frequency range of 0.0018 Hz ~ 20 kHz were
obtained on each transect. The remote reference technique (Gamble et al., 1979) was adopted in order to minimize effect of noise. Observations at two sites, separated by about 70 km, were conducted simultaneously, and one site was treated as the reference site for the other. Additional survey on three short transects were conducted north of Line A in order to investigate axial variation of the resistivity distribution in the Central Mountain Range.

Data quality is generally good in the mountainous area, but not in urban areas along the eastern heavily populated plain area. Several steps to improve data quality were taken besides the remote reference technique, such as smoothing of the impedance tensor in the frequency domain through the 1-D Bostic model. The smoothed impedance data are sampled at chosen frequencies on the basis of the 1-D model. Beside extraneous noises in the MT observation, we need to correct the data for near surface 3-D heterogeneity to get reasonable 2-D model. The electrical resistivity structure in the survey region was shown to be nearly two-dimensional (Fujinawa et al., 1997; Kawakami et al., 1997). The effects of subsurface 3-D heterogeneities have been quantitatively estimated by means of the Groom-Bailey tensor decomposition (GBD) method (Groom and Bailey, 1989, 1991; Groom et al., 1993). This practical technique provides corrected two-dimensional impedance $Z_{2D}$ from the observed impedance tensor $Z_{ob}$. The GBD decomposition procedure provides values for regional strike direction $\phi$, thrust $t$, and shear $e$, but not site gain $g$. The site gain was determined using the result of the TEM.

The GBD method indicated that the 3-D heterogeneity effect at major sites could be reasonably described as a 3-D galvanic distortion over the 2-D inductive structure. The estimated regional strike direction was found to align roughly along north-northwest to south-southwest, though with considerable scatter at several sites. The strike correlates well with that of the geological units and the axis of the island arc, though little bit inclined toward west (Kawakami et al., 1997). At a few sites, the GBD result show that the 3-D effects are not consistent with the assumed 3-D galvanic distortions. Such sites are marked by large skew angles. At such sites, the GBD indicated structural trend of the Neogene units are nearly parallel to the trench axis but different from the trend of the pre-Tertiary basement rock units (Kimura et al., 1991). Results of 2-D modeling should be treated carefully beneath these sites particularly at depths of the lower crust and upper mantle.

In the tensor distortion analysis, the site gain factor related to the static-shift cannot be calculated, and so needs to be inferred by some other means. In this study, a time-domain electromagnetic (TEM) measurement method (Sternberg et al., 1988) was applied to handle MT static-shifts (Jiracek, 1990), utilizing a TEM-FAST ProSystem (AEMR Co., Ltd.). Coincident in-loop TEM measurements were conducted using a rectangular antenna configuration with total length 200–300 m (Fujinawa et al., 1999). One-dimensional analysis of the TEM data was used to obtain layered models at the shallow depths. Frequency distribution of the TEM apparent resistivity and phase were merged with the MT measurements for both the orthogonal TE and TM modes. The amount of shifts turned out to be less
than 0.2 decade for 70% of the data. The simple mean of the absolute value is 0.084 decade, and the standard deviation $\sigma = 0.24$ decade. Some sites, however, exhibit large shifts of the order of a decade or more, indicating that the static shift corrections should be applied prior to modeling.

3. 2-D MODELING

In spite of the considerable scatter in the ensemble of strike directions in the survey area, we assume, for simplicity, that the regional strike in the 2-D modeling is N-S (Fujinawa et al., 1997, 1999). The 2-D MT inversion algorithm, GRRI (Lee et al., 1995; Yamane et al., 1996), was applied to this data. The algorithm is an extension of the RRI algorithm developed by Smith and Booker (1991). Effects of the Pacific Ocean and the Japan Sea are simulated by assuming 50 m thick surface layer with resistivity of 0.2 ohm-m (salt water) extending out from the coast.

The final 2-D models along the three traverses are presented in Fig. 3. The earthquake hypocenter, deduced through the seismic network of Tohoku University (Hasegawa et al., 1991), is overlain for use in later discussion. The convergence in inversion is judged rather subjectively, based on degree of fit of the model data to the observed data as well as on overall smoothness of the model. Generally, there is a good fit between the model and the data, though there are several sites with a large degree of misfit over certain frequency bands (Fujinawa et al., 1997, 1999).

Here, the TM-mode models are displayed though we obtained the 2-D models both for the TM, TE as well as for joint modes. The TM mode inversions resulted in models that fit the observed data at almost all sites along the three traverses over the specified frequency range. We can expect that the TM mode solution can image the resistivity profile sufficiently well in the presence of the 3-D heterogeneity (Wannamaker et al., 1984). Inverting the TE mode, however, resulted in a quite different model compared with the TM model, though the fit of the TE model to the data was satisfactory at most of sites. We could not succeed in getting the TE model consistent with that obtained for the TM mode, in spite of many trials with different mesh design, deletion of neighboring sites with large differences (contrasts) in apparent resistivity, or supplementing the dataset at additional sites near the problematic locations. Therefore, the TE modeling is not treated in this paper and would form the subject of future investigation.

Conductive structure along Line A

We can see two prominent conductive bodies, CB around the Central Basin (around sites 605–203), and KK, east of the Central Mountain Range around the Kitakami River (around the sites 306–612). A localized conductive anomaly appeared in the Central Mountain Ranges (SE) under site 204, which has little similarity with surrounding sites in terms of 1-D resistivity model. The anomaly is not considered to be artifact of numerical modeling, but a local heterogeneity related to geothermal activity at a nearby Quaternary volcano. In order to test this
Fig. 3. Two-dimensional resistivity models for the TM mode along the three transects using the corrected MT impedance tensors. Earthquake hypocenters determined by Tohoku University (Hasegawa et al., 1991) are overlain with active faults (Fi*) and geological lineaments (Fi).
hypothesis, supplementary MT data were acquired at site 902 in the later phase of the experiment, which is just east of site 204 (Fig. 1). The apparent resistivity profile at site 902 turned out to be very similar to those at sites 304 and 609 having no prominent conductive anomaly. Therefore, the data at site 204 was discarded because the present interest is to find a regional 2-D modeling.

Another conductive anomaly (PC) was located near the Pacific coast, around sites 308 and 309. In order to resolve the problem, additional data were acquired at site 903 near the conductive anomaly PC. The additional data does not indicate the strong conductor in this region, though the presence of a weak conductive body cannot be ruled out. Therefore, we assume that the PC anomaly is related with possible 3-D effects. A weakly conductive body under sites 611 and 612 near KK is poorly resolved. However, we can assume that there is a significant heterogeneity, because the apparent conductor corresponds to a singular zone of most active seismicity, as it will be discussed later.

A conductor located near the Japan Sea (JP) under site 602 was checked by supplementary observation at sites 901 and 907 located between sites 601 and 603. The 1-D models for the supplementary dataset indicated that the region around site 901 is nearly the same as that of its immediate easterly site 603. Site 907 is found to be in a transition zone between the neighboring regions. Search for suitable model suggested that the vertical stripping (seemingly excessive extension of conductivity feature downward) may be caused by either large 3-D heterogeneities, or it may be simply artifact of inappropriate mesh design. Using a finer mesh in the near-surface modeling resulted in a conductor with vertical extent of less than 1 km. This is several times smaller compared with the vertical extent of the JP anomaly at site 602 (Fig. 3). We could assume that there is a conductive body in the upper crust and possibly another conductive body near the Conrad Discontinuity under site 602.

Conductive structure along Line B

On Line B, there are two prominent conductors: a thin, extended surficial conductor, east of the Central Mountain Range between sites 408 and 412, and a very prominent conductor west of the Central Mountain Range (CB) between sites 402 and 404. The conductor (CB) probably corresponds to an extensive fracture zone west of the Yamagata Basin, as presented in Figs. 1 and 3. The upper part of the Yamagata basin is seen as a thin conductor with a thickness of less than 1 km (Fig. 3). Similarly thin conductor can also be seen under site 712. This may raise some doubts that the large vertical extent of the western conductive anomaly CB may be affected by numerical artifact. But it would be probable that large vertical extent of the conductive body is brought out by a group of stations and, thus, may denote major fault, extending deeper than the Conrad Discontinuity. We note the similarity with the deep penetration of a fracture zone found in western Quebec (Calvert et al., 1995; Tournerie and Chouteau, 1998). But caution is needed in making this comparison because of the different tectonic regimes of the two regions.
A resistive region in the upper crust, west of the conductive fracture zone, is thought to correspond to the plutonic rocks around Mt. Asahi. We note that the lower crust in this region is relatively more resistive than the upper crust.

**Conductive structure along Line C**

The model presented for Line C (Fig. 3) indicate a thin but clear conductive body east of the Central Mountain Range and a very prominent conductor west of the Central Mountain Range extending down to Moho. The conductor is much smaller in the eastern half of the Central Basin. The conductor CB is considered to represent a fracture or fault zones in the region between sites 704 and 706.

The vertical-stripping under the conductor east of the pre-Tertiary Hatagawa Tectonic Line (HTL) appeared in the model using uncorrected impedance. But the stripping has almost disappeared in the corrected model. This may be due to the distortion corrections made to the data at site 906. In this model, the conductive sediment corresponding to the Yamagata basin is very shallow compared with that of the fracture zone.

4. **ELECTRICAL CHARACTER OF CONTINENTAL LOWER CRUST (CLC)**

The lower continental crust is generally characterized by an enhanced conductivity, though there may be a considerable degree of variability (Haak and Hutton, 1986; Jones, 1992; Simpson, 1998). The fluid-rich conducting layer in the CLC has been considered to be a dominant feature in several geodynamic models, and has been often used to explain aseismic lower crust (Schmeling and Marquart, 1990; Kaufman and Royden, 1994). The lower crust in the present survey area is shown to be seismically inactive (Hasegawa _et al._, 1991), indicating that the lower crust is ductile. It is interesting to examine the situation through the conductivity distribution data, obtained here.

On Line A, conductors are primarily confined to the upper crust. The resistivity does not seem to decrease downward from the upper crust to the lower crust, but rather increases with depth down to mantle depths of at least 100 km. The crustal resistivity structure along Line B is similar to that along Line A. The 1-D models also suggest an almost homogeneous resistivity distribution in the whole crust and upper mantle including region of Dewa Hill (characterized by a pluton) and the Central Mountain Range. The resistivity distribution actually increases with depth under the surficial conductor KK.

On Line C, except for the two local conductive bodies CB and KK, entire crustal cross-section is generally resistive with resistivities of several hundreds ohm-m. The present results indicate the resistivity of the order of 1,000 ohm-m beneath Dewa Hill (west of CB), the Central Mountain Range and under the conductive body near the Pacific coast. At depth, the resistivity on both sides of CB appears to be smaller in comparison with the upper part. This may be, however, due to defects in the model as is indicated from the larger misfit of the apparent resistivity curves.
In summary, the lower crust in the region cannot be claimed to be more conductive than the upper crust. It is probably uniform over the whole crust, or the lower crust may even be more resistive in the central part of the northeastern Japan arc, as compared to the northern and southern parts (Ogawa, 1992; Ogawa et al., 1992). The lower crust in the northeastern Japan arc may not be regionally uniform as indicated by Honkura (1988). Our models show that the resistivity in the lower crust varies from several hundreds ohm-m to thousand ohm-m. Haak and Hutton (1986) have classified this range of resistivity as “normal” for the lower crust.

The enhanced conductivity of LC, reported in the various tectonic situations, is explained by the presence of saline fluids, black shales and/or graphite, and partial melting (Jones, 1992; Simpson, 1998). The partial melt hypothesis (Chen et al., 1996), however, could not be applicable to the present study area, because no pronounce seismic S-wave reflectors, which suggest existence of fluid phase substance, are seen in the survey area (Umino and Hasegawa, 2002). Saline fluids enhance the conductivity in the premise of an interconnected network. The fluid could be trapped below the impermeable layer at the Conrad Discontinuity (Jones, 1987) resulting in a ductile lower crust without appreciable seismic activity.

The absence of enhanced conducting layer in the present area or the result that upper crust is more conducting than lower crust can be explained if the above trapped fluid mechanism beneath the impermeable layer (Jones, 1987; Hyndman, 1988) is not effective in tectonic scenario of island arc. If the saline fluids are assumed to be insufficiently sealed in the mid-crust in the case of the island arc, the lower crust will not be seen as more conducting that upper crust or even upper crust may tend to be more conducting due to the upward propagation of fluids from lower crustal depths. Yokokura et al. (1998) provided evidence of a breakdown of the sealing in the back-arc region on the basis of seismically transparent Conrad and Moho Discontinuities in these regions.

5. RELATIONSHIP BETWEEN SEISMICITY AND CONDUCTIVITY DISTRIBUTION

Figure 3 shows the resistivity model overlain with a vertical cross-section of hypocenters of earthquake within 20 km wide corridor centered on each traverse. The data are from the seismic database at Tohoku University (Hasegawa et al., 1991). Figure 4 shows the spatial distribution of the hypocenters of earthquakes in the upper crust with focal depths from 0 to 15 km against the background of the resistivity distribution in the upper crust (at a depth of 7.5 km). Historical earthquakes are indicated by star symbols (★). Here, the horizontal resistivity distribution has been obtained through the spline interpolation of the three vertical profiles by assuming a quasi-2-D structure. Consequently, the correlation between conductivity anomalies and tectonic units can be discussed on a spatial scale of the order of the distance between neighbouring sites. In Figs. 3 and 4 the pre-Tertiary tectonic lines, Tanakura (TTL), Hatagawa (HTL), and Morioka-Shirakawa (MSL) are included with the Quaternary Volcanic Front (VF) (see Fujinawa et al., 1997 and references therein for more detail).
Fig. 4. Horizontal resistivity distribution at a depth of 7.5 km, within the upper crust, and is overlain by earthquake hypocenters as determined by Tohoku University's dense seismic network and major historical earthquakes. Solid curves attached with number F* indicate active faults (Active Fault Research Group, 1991). The solid curve with the mark F indicates active faults or lineaments (Geological Survey of Japan, 1995).

The prominent conductive bodies in the back-arc side are seen to be well correlated with the active faults (F*) and geological lineaments (F) located in this region. These include a fracture zone west of the Central Basin, the Shinjo basin (around sites 606–203 on Line A), the Yamagata basin around sites 405–406 on Line B, and around sites 904–506 on Line C. Seismicity is very high in these regions indicating that the stress is being released in the fracture zones. A fluid-rich regime in the fracture zone may weaken the crust enough to trigger earthquakes. This inference will be reinforced further when the Poisson ratio data is used in later discussion.

The seismicity near the resistive Dewa Hill, west of the Central Basin, however, is very low in comparison with that observed in the fracture zones in spite of many small faults. We know that the conductors do not always correspond to active faults. The seismicity is controlled by many factors: stress accumulation, fault strength and regional tectonic regime. A candidate for explaining the low seismicity in this region may be the absence of water. In the Central Mountain Range, we can assume that there is a different seismo-tectonics situation compared with the Dewa Hill. Narrow zones, especially around the volcanic front (VF), are seismically active while the region seems to be dry, marked by high resistivity.
There are several isolated spots of high seismicity in Fig. 4. We can see two moderate earthquakes occurred along the eastern edge of the conductor corresponding to Central Basin, i.e. around site 405 on Line B and, on Line C, near site 904. These epicenters seem to be aligned along the line that connects to the active fault F7* (Funagata Fault; Active Fault Research Group, 1991), passing west of site 203 in the north and continuing along the northern tip of another geological lineament located far south of site 706 (see Fig. 1; Geological Survey of Japan, 1995). This line appears to demarcate sharp electrical boundary with conductive block to the west and highly resistive block to east. This line probably corresponds to a geological boundary separating the resistive Central Mountain Range from the conductive Central Basin. The line has not been identified earlier on the active fault map (Active Fault Research Group, 1991).

Several kilometers northwest of site 710 there is a region with concentrated seismicity, and a moderate earthquake with a magnitude of 5.8 occurred in 1706 nearby indicating an active fault. No anomalous conductivity, however, was found in our work. Moreover, there are no traces of active faults or geological lineaments reported. The region is difficult, however, to approach for MT measurements, and there are no means to investigate the curious seismicity there waiting for another approach.

The region around site 612 on traverse A, about 20 km south of site 206 is very active seismically. Moderate earthquakes with a magnitude of 7.0 and of 6.5 (Miyagi-ken-hokubu Earthquake) occurred in 1900 and 1962, respectively. The region is characterized as the eastern edge of the KK conductor (Fig. 3) and is on the Shirakawa-Morioka Line (Figs. 3 and 5). The relation between the seismicity

Fig. 5. Hypocentral distribution of earthquakes in the lower crust between depths of 15 to 30 km, plotted on the modeled resistivity distribution at a depth of 22.5 km within the lower crust.
(including micro-earthquake activity) and electrical conductivity is not clear, suggesting that other parameters are involved in the earthquake occurrence process, depending on the tectonic situation.

It is now widely accepted that the lower crust is ductile. The close agreement between the depths to the conductive lower crust, seismic depth cut-off in the crust as well as the brittle-ductile transition are considered strong pointer of the onset of ductile layer (Meissner and Strehlau, 1982; Stanley, 1989; Adam, 1978; Hyndman and Shearer, 1989; Simpson, 1998). The Curie point depths, estimated from the geomagnetic survey data (Okubo et al., 1985), are 10 ± 2 km in the survey area, indicating that the lower crust is in ductile condition. Spatial distribution of the brittle-ductile transition depth was obtained using precisely located micro-earthquakes hypocenter and the temperature distribution calculated from the P wave velocity in the Central Mountain (Umino and Hasegawa, 2002). From this point of view, the resistivity data are expected to provide better estimate of the temperature than that has been based on the seismic velocity data (Sato, H. et al., 1989).

Figure 5, similar to Fig. 4, gives spatial distribution of hypocenters (located between 15–30 km) as well as resistivity distribution around a mean depth of 22.5 km. It is seen that the lower crust is predominantly aseismic (Fig. 5). There are, however, some of scattered earthquakes of low frequency (Umino and Hasegawa, 2002). The seismicity is extremely low in the highly resistive region. A small pocket of high seismicity is observed near the geothermal area around sites 204 and 305. These sites are near the Quaternary volcanoes; Komagatake and Onikobe. This had led Umino and Hasegawa (2002) to conclude that these low-frequency earthquakes are related to magma activity.

6. CORRELATION OF SEISMICITY WITH MULTIPLE GEOPHYSICAL DATA

In addition to the information on conductivity distribution and active faults, the seismic velocity data have been used to provide insight on the complex relation between particular seismic activity and crustal regime (Zhao et al., 1992). The P-wave velocity (Vp) and S-wave velocity (Vs) are calculated using the regression formula to represent velocity values at grid nodes for the upper crust \( (H = 7.5 \text{ km}) \) and for the lower crust \( (H = 22.5 \text{ km}) \). The Poisson ratio \( \sigma \) is deduced from the Vp and Vs values. Several interesting correlations between the seismicity and these new geophysical data sets emerge in various zones. In Table 1, we have summarized the features of geological and geophysical information at these zones for the sake of convenience.

The Asahi Mountain Range with many of active faults is contrasted from the Central Mountain Range, in the sense that former is seismically much less active than later. In terms of resistivity distribution, both are marked by nearly equal high resistivity (Table 1(a)). The other striking difference is that relatively low values of the Poisson ratio (~0.20–0.22) prevail beneath the Central Mountain Range compared with the Asahi Mountain Range. This can be construed to be the pointer of relatively poor saturation (fluid) state of rocks beneath the Central
Table 1(a). Characteristics of geophysical and geological parameters in the upper crust at the typical zone in terms of seismicity.

**Upper Crust**

<table>
<thead>
<tr>
<th>Index Zone</th>
<th>( \rho ) (O·m) (5 \sim 1,000)</th>
<th>(V_p(\text{Km/s})) (5.5 \sim 6.2) Av. = 5.85</th>
<th>(V_s(\text{Km/s})) (3.2 \sim 3.7) Av. = 3.45</th>
<th>(\sigma) (0.20 \sim 0.30) Av. = 0.25</th>
<th>Fault</th>
<th>Tectonic Line</th>
<th>Seismicity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mt. Asashi</td>
<td>( &gt; 500 )</td>
<td>+</td>
<td>(6)</td>
<td>(3.5)</td>
<td>(0.24)</td>
<td>Many</td>
<td>TTL</td>
</tr>
<tr>
<td>Fracture Zone</td>
<td>(5)</td>
<td>-</td>
<td>(5.9 \sim 6.0)</td>
<td>(3.6)</td>
<td>(0.20)</td>
<td>Many</td>
<td>TTL</td>
</tr>
<tr>
<td>Central Basin</td>
<td>(5)</td>
<td>-</td>
<td>(5.9 \sim 6.9)</td>
<td>(3.6)</td>
<td>(0.20 \sim 0.22)</td>
<td>Many</td>
<td>None</td>
</tr>
<tr>
<td>Central Mountain Range</td>
<td>(1,000) (north:100)</td>
<td>+</td>
<td>(6.0)</td>
<td>(3.5 \sim 3.7) +</td>
<td>(0.20 \sim 0.22)</td>
<td>Two</td>
<td>VF HTTL</td>
</tr>
<tr>
<td>Miyagi-ken- Hokubu</td>
<td>(30)</td>
<td>-</td>
<td>(6.2)</td>
<td>(3.5)</td>
<td>(0.25)</td>
<td>None</td>
<td>MSL</td>
</tr>
</tbody>
</table>

Table 1(b). Same as Table 1(a) except in the lower crust.

**Lower Crust**

<table>
<thead>
<tr>
<th>Index Zone</th>
<th>( \rho ) (O·m) (5 \sim 1,000)</th>
<th>(V_p(\text{Km/s})) (6.3 \sim 7.0) Av. = 6.6</th>
<th>(V_s(\text{Km/s})) (3.4 \sim 4.1) Av. = 3.75</th>
<th>(\sigma) (0.2 \sim 0.3) Av. = 0.25</th>
<th>Fault</th>
<th>Tectonic Line</th>
<th>Seismicity</th>
</tr>
</thead>
<tbody>
<tr>
<td>Mt. Asashi</td>
<td>(100 \sim 200)</td>
<td>+</td>
<td>(6.4 \sim 6.5)</td>
<td>(3.6)</td>
<td>(0.25 \sim 0.26)</td>
<td>Many</td>
<td>TTL</td>
</tr>
<tr>
<td>Fracture Zone</td>
<td>(5 \sim 10)</td>
<td>-</td>
<td>(6.5 \sim 6.6)</td>
<td>(3.6 \sim 3.8)</td>
<td>(0.27)</td>
<td>Many</td>
<td>TTL</td>
</tr>
<tr>
<td>Central Basin</td>
<td>(20 \sim 50)</td>
<td>-</td>
<td>(6.6 \sim 6.7)</td>
<td>(3.6 \sim 3.8)</td>
<td>(0.27)</td>
<td>Many</td>
<td>None</td>
</tr>
<tr>
<td>Central Mountain Range</td>
<td>(500 \sim 1000)</td>
<td>+</td>
<td>(6.6 \sim 6.7)</td>
<td>(3.6 \sim 3.8)</td>
<td>(0.26)</td>
<td>Two</td>
<td>VF HTTL</td>
</tr>
<tr>
<td>Miyagi-ken- Hokubu</td>
<td>(2000) c.f. Upper Crust</td>
<td>+</td>
<td>(7.0) c.f. Upper Crust</td>
<td>(3.9 \sim 4.0)</td>
<td>(0.28)</td>
<td>None</td>
<td>MSL</td>
</tr>
</tbody>
</table>

Mountain Range. The fluid poor state in the Central Mountain Range may be related with the volcanism (Yokokura et al., 1998). The inference is in accord with the absence of enhanced conductivity in the lower crust (Fujinawa et al., 1999). It may be argued that seismicity, particularly micro-earthquakes, in the brittle crust of this region may be activated by the tectonic stress rather than the physical state of medium. The situation may be similar to the upper crust in the rupture zone of Kobe-earthquake (Zhao and Negishi, 1998).
The fracture zone in the west of the Central Basin shows exceedingly small value of the Poisson ratio of 0.20 that varies in the interval 0.20–0.30. Here, the seismic velocity Vs is large, same is the case beneath the Central Mountain Range. Another common feature in two regions is that there are no big earthquakes; i.e. strains cannot be accumulated enough to trigger a larger scale rupture.

In the Miyagi-ken-hokubu region, large earthquakes have occurred twice in this century. The zone is characterized as conductive and has local maximum of Poisson ratio, suggesting relatively fluid-rich state. A situation noted with the main-shock zone of the Kobe-earthquake (Zhao and Negishi, 1998). Moreover, the Vp is found to be largest in the area indicating very brittle nature of the crust.

The above examples show that use of the seismic velocity and Poisson ratio can be helpful in the interpretation of seismicity by taking into account of state of rock including degree of fluid content. However, the various parameters appear to combine in complex manner (Table 1) to produce numerous patterns of seismotectonics, with a result that no definitive inference can be drawn with presently available data. Higher spatial resolution in the conductivity, seismic velocity, Poisson ratio would be very useful in bringing clear picture of the seismotectonics in the given area.

The low values of the Poisson ratio in the Central Basin and Central Mountain Range does not extend to the lower crust (Table 1(b)) suggesting clearly different character of the lower compared with upper crust, in conformity with the difference in seismicity. On the other hand, we note that the conductivity and the Vp are both anomalous in the whole crust in the Miyagi-ken-hokubu earthquake zone. It is suggested that the entire crust is anomalous, indicating a large scale of heterogeneity there.

7. CONCLUSIONS

Wide-band MT observations have been conducted along six transects in the central part of the northeastern Japan arc. The magnetotelluric impedance tensors were corrected for possible 3-D distortion effects including static-shifts by using TEM data. 2-D models were obtained using the GRII inversion algorithm to build a quasi-three dimensional resistivity distribution. The model shows that:

1) The lower crust is not particularly conductive compared with upper crust.
2) In the major fracture zones at the west of the Central Basin, there is a good correlation between conductive bodies, faults or geological lineaments and seismicity.
3) The Central Mountain Range and Dewa Hill are characterized, both by highly resistive bodies and low Poisson ratio, suggesting fluid-poor state.
4) Active seismicity around the boundary of the Central Basin and the Central Mountain Range may be due to a buried fault, appearing as a sharp conductivity discontinuity.
5) Many of the faults associated with scarce seismicity are characterized by higher resistivities.
6) Tectonic stress level is not accumulated in the Asahi Mountain Range.
Acknowledgements. The authors express their sincere thanks to Drs. Hiroshi Takahashi, Shigetsugu Uyehara, Gensuke Endo (late), and Prof. Sachio Nabetani for their constant encouragement. Thanks are also expressed to Profs. Nobuaki Nizuma, and Yoshihiro Suzuki for valuable discussions and suggestions, and to the many organizations and institutes that helped us to conduct experiments. We thank greatly Prof. Zhao Da Peng providing programs to calculate seismic velocity. Many valuable comments by Prof. Baldev K. Arora are greatly appreciated. I was helped to prepare the manuscript by Ms. Yumiko Yamauchi and Mr. Wu Yuesheng. The Science and Technology Agency Japan funded the study for the Promotion of Surveys and Research on Earth Science and Technology and Ocean Development.

REFERENCES


Horiuchi, S., H. Ishii and A. Takagi, Two-dimensional depth structure of the crust beneath the Tohoku District, the northeastern Japan arc, I. Method and Conrad discontinuity, *J. Phys. Earth*


